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Review

The age and origin of the Pacific islands: a geological overview

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The Pacific Ocean evolved from the Panthalassic Ocean that was first formed *ca* 750 Ma with the rifting apart of Rodinia. By 160 Ma, the first ocean floor ascribed to the current Pacific plate was produced to the west of a spreading centre in the central Pacific, ultimately growing to become the largest oceanic plate on the Earth. The current Nazca, Cocos and Juan de Fuca (Gorda) plates were initially one plate, produced to the east of the original spreading centre before becoming split into three. The islands of the Pacific have originated as: linear chains of volcanic islands on the above plates either by mantle plume or propagating fracture origin, atolls, uplifted coralline reefs, fragments of continental crust, obducted portions of adjoining lithospheric plates and islands resulting from subduction along convergent plate margins. Out of the 11 linear volcanic chains identified, each is briefly described and its history summarized. The geology of 10 exemplar archipelagos (Japan, Izu-Bonin, Palau, Solomons, Fiji, New Caledonia, New Zealand, Society, Galápagos and Hawaii) is then discussed in detail.

Keywords: Pacific; islands; plate tectonics

1. INTRODUCTION

The islands of the Pacific Ocean support biodiversity that has evolved in the context of diverse and complex geological histories. Understanding the evolution of this biota requires an understanding of the geological history of the Ocean and its many thousands of islands. Two-thirds of the Ocean is underlain by the large northwestward-moving Pacific plate that is generated at the East Pacific Rise (EPR) and its southern continuation, the Pacific–Antarctic Ridge (figure 1).[1] The plate extends approximately 13 000 km from source to subduction[2] in the Mariana Trench, moving at 50–133 mm yr^{-1} about a pole of rotation located on the east Australian coast (Wessel & Kroenke 2000). Thus, in its western part, the Pacific plate preserves the oldest ocean floor on Earth at 167 Ma, of Jurassic age (Koppers *et al.* 2003).

The Pacific plate comprises solid lithosphere[3] that moves above a weak ductile layer in the upper mantle, named the asthenosphere.[4] Lithospheric plates comprise an upper layer of crustal rocks beneath which is a higher density layer of solid upper mantle. The crust and upper mantle of each plate behave coherently without movement between them. The crust is made of either dense basaltic lavas and their intrusive equivalents, termed oceanic crust, or lower density sedimentary rocks, granites and their extrusive and metamorphosed equivalents, termed continental crust. Some lithospheric plates are made entirely of oceanic or continental crust, but the vast majority comprise mosaics of both crust types, depending on their previous geological histories.

Oceanic crust is thin close to its source at mid-ocean spreading centres, and thickens to 10 km in distal locations. By contrast, continental crust varies from 20 to more than 70 km thickness. Where continental crust converges with oceanic crust, that part of the lithospheric plate with the denser (oceanic) crust slips downwards and sinks into the asthenosphere, forming a subduction zone. Earthquakes created by the fracturing of the cold, brittle subducting lithosphere can extend to depths of 700 km. Water released from the upper layers of the descending plate interacts with the surrounding asthenosphere (the mantle wedge), eventually creating magmas that rise to the surface forming volcanic arcs. Intermixing of the source magma with assimilated crustal materials on its upward journey creates a common volcanic rock type named andesite. A line of andesite volcanoes (‘the andesite line’) circumscribes the Pacific Ocean clockwise from New Zealand to Chile, and is colloquially known as ‘the Pacific Ring of Fire’. Where oceanic crust converges with oceanic crust, it is the older, denser plate that will subduct, such as that in the Tonga–Kermadec arc. Where continental crust converges with continental crust, neither can be subducted. The result is crustal overthickening, compressed rock sequences, thrust faulting and mountain building, termed obduction.[5]

On the Pacific plate’s journey from source to subduction, the lithosphere becomes progressively colder, and denser, leading to gradual subsidence over time. East of the EPR, three secondary plates (Juan de Fuca (Gorda), Cocos and Nazca) form most of the rest of the sea floor of the Pacific Ocean. The first

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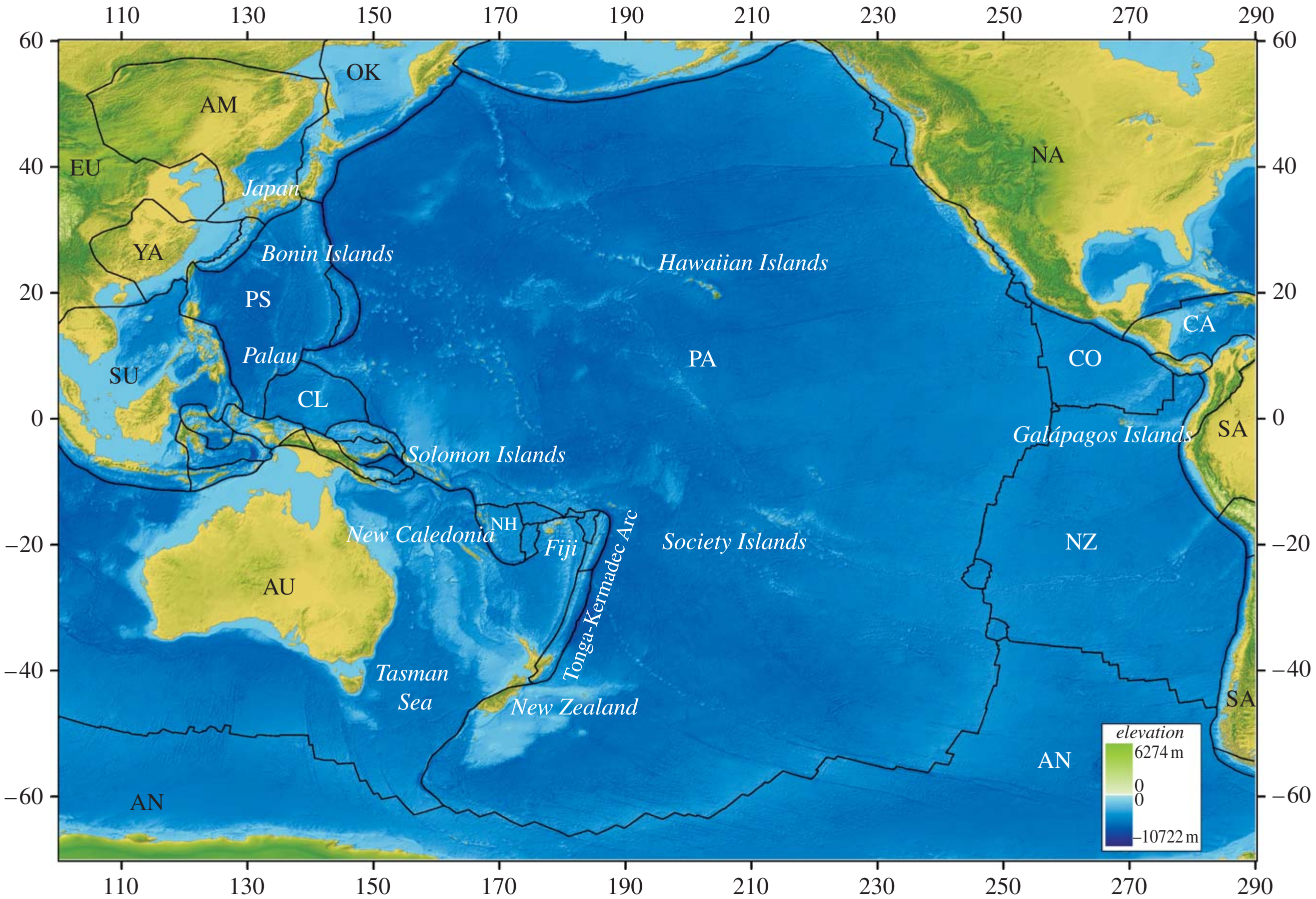

Figure 1. Map of the Pacific Ocean showing the location of islands and major geological features discussed in the text including major plates (according to Bird 2003). Plate nomenclature symbols are: AM, Amur; AN, Antarctic; AU, Australian; CA, Caribbean; CL, Caroline; CO, Cocos; EU, Eurasia; NA, North American; NH, New Hebrides; NZ, Nazca; OK, Okhotsk; PA, Pacific; PS, Philippine Sea; SA, South American; SU, Sunda; YA, Yangtze. For microplate nomenclature, see the electronic supplementary material.

two move northeastwards and the latter eastwards, to be subducted beneath North, Central and South America. To the south of the Pacific–Antarctic Ridge is the Antarctic plate to which new oceanic crust is being added on almost all its boundaries. In the west and southwest Pacific, an array of microplates buffers movement between the Pacific and the Australian and Eurasian plates, from New Zealand in the south to Taiwan in the north. This has yielded a complex series of short-lived mid-ocean spreading centres, plate rotations and concomitant subduction.

Due primarily to plate creation at the EPR, a large part of the Pacific Ocean margin consists of a series of subduction zones where the Pacific, Cocos and Nazca plates plunge into the Earth's asthenosphere. Associated andesitic volcanism along most of the western coastline of the Americas has occurred on the adjoining continental lithosphere, but from the Aleutians to New Zealand in the western Pacific it produced island arcs.[6] Exceptions are where plates move horizontally past each other, resulting in strike–slip faulting,[7] such as along the San Andreas Fault in California and the Alpine Fault in New Zealand.

2. ANCIENT ORIGINS OF THE PACIFIC OCEAN

The origins of what now comprises the Pacific Ocean extend back 750 Ma to the rifting of the Proterozoic continent, Rodinia. This continent split into two creating the Panthalassic Ocean, the ancient predecessor to the Pacific Ocean. This may have been triggered by the appearance of a Pacific super plume[8] (Taira 2001). By 400 Ma, the supercontinent of Pangea began to form, centred on the Equator and extending from the North to the South Poles. It separated the Palaeo-Tethys Ocean to the east from the Panthalassic Ocean to the west.

By 180 Ma the first signs of break-up of Pangea were heralded by rifting that first formed the central Atlantic Ocean. By 167 Ma, sea floor spreading to the west of Pangea formed ocean floor that we see today as the oldest ocean floor of the Pacific plate (Koppers *et al.* 2003). This new sea floor emanated from a Y-shaped ancestral EPR and its extensions, creating three major oceanic plates named clockwise from the north, the Kula, Farallon and Pacific. Subsequently, the spreading ridges forming the V of the Y moved northwards and eastwards to be subducted beneath the adjoining continental lithosphere of the North American plate.

By 140 Ma, Gondwana (the southern portion of Pangea) had begun to fragment leading to India separating from Madagascar to form the Indian Ocean. The western coastline of South America is now a zone of convergence with the Pacific Ocean, as was the contiguous coast of east Australia–Antarctica. The Tethys Ocean separated eastern Gondwana from eastern Laurasia.

At 50 Ma, the Earth was beginning to attain its current form and the arrangement of oceans and continents with which we are familiar with today. By this time, India had begun to collide with Asia, closing the Tethys Ocean and beginning the formation of the Himalayan Mountains and the Tibetan Plateau. At *ca* 33.5 Ma, Australia finally separated from Antarctica and moved northwards to its present position (Exon *et al*. 2001). Over this time, the Pacific plate expanded to finally occupy two-thirds of the current Pacific Ocean. East of the EPR, the Farallon plate moved eastwards until a portion of the spreading ridge reached California. This represented the first contact of the Pacific plate with the North American plate, and relative movement between them was predominantly transcurrent (strike–slip) in the form of the San Andreas Fault. The fragment of the Farallon plate to the north became the Juan de Fuca or Gorda plate, while the fragment to the south broke into two, forming the Cocos and Nazca plates. The Kula plate to the north was subducted northwards beneath Alaska and no longer existed by 20 Ma. For a more detailed account of these relative plate movements, see the Paleomap Project site at www.scotese.com.

Palaeomagnetic stratigraphy and deep-sea drill cores support this model by demonstrating the progressive increasing age of sea floor away from the EPR. The fast spreading rate means there was insufficient time for volcanic islands to be built along the EPR axis, so all the volcanics are submarine. This contrasts with slower sea-floor spreading centres in other oceans, such as the Mid-Atlantic Ridge that rises above sea level in Iceland.

3. ISLAND ORIGINS

We distinguish five major processes of island formation in the Pacific as follows.

(a) Formation of volcanic island chains and seamounts. Wilson (1963) first proposed that islands of the Hawaiian chain could be explained by lithosphere moving over a stationary hot spot in the mantle. Each island could have formed as basaltic magmas rose through the lithosphere to the surface, forming active large shield volcanoes. As the plates moved, so alignments of extinct volcanic islands record the direction and timing of this movement with respect to a reference frame of stationary hot spots (see the electronic supplementary material). Morgan (1971, 1972) proposed that hot spots were continually supplied by a plume from the deep mantle. This explanation is now being challenged and a number of alternative hypotheses proposed. One suggests that volcanic island chains are formed along a pre-existing or propagating fracture in a lithospheric plate due to stresses transmitted from the plate margins (Smith 2003, 2004; Natland & Winterer 2005). The en echelon[9] arrangement of volcanic ridges along the Pukapuka Seamount Chain (Winterer & Sandwell 1987; Sandwell *et al*. 1995) supports this tensional origin. The volcanic chain of the Line Islands also shows no progressive ages along its length, supporting a synchronous *ca* 8-Ma origin consistent with a mechanism of lithospheric extension (Davis *et al*. 2002). Another is that the volcanic chains might be caused by warps and cracks caused by uneven thermal contraction of the cooling lithosphere (Sandwell & Fiialko 2004; Ballmer *et al*. 2007). Courtillot *et al*. (2003) suggested that hot spots may result from three separate mechanisms (i) a deep lower mantle/outer core source, (ii) the transition zone between the upper and lower mantles or (iii) from within the upper mantle immediately beneath the lithosphere.

(b) As volcanic islands move away from their birthplace, so the oceanic crust becomes progressively colder and denser, subsiding from the level of its formation. In this manner, each volcano gradually sinks as it moves farther from its point of origin. In tropical waters coral growth forms fringing reefs, and further subsidence leads to the formation of atolls built on the underlying subsided volcanic edifices, as first postulated by Darwin (1890). These stages are superbly displayed in the Cook Islands between Rarotonga (fringing reef), Aitutaki (atoll with emergent volcanic peak at northern end) and Palmerston (complete coral atoll). Eventually, the volcanic edifice sinks below sea level to progressively greater depths to form a seamount or guyot.

(c) Flexing of the lithospheric plate in some regions may lead to uplift. In some circumstances, this is due to seamount loading of the oceanic lithosphere creating an annular moat and surrounding arch (Lambeck 1981). At scattered locations throughout the South Pacific, coral reefs that once grew in the marine environment above an extinct volcano have been uplifted above sea level. Subsequent erosion by the sea has resulted in steep coral cliffs circumscribing the islands, colloquially known as 'Makatea'. Such islands are composed entirely of coralline limestone, and classic examples include Niue (maximum height of 65 m), Henderson Island (at 30 m) and Makatea (113 m). Rennell Island in the Solomon archipelago is the largest raised coral atoll in the world, being surrounded by 120–150 m high cliffs.

(d) Owing to past and present plate movements, fragments of continental crust have been rotated away from nearby land masses to now occupy isolated positions in the Pacific. The largest of these is Zealandia (Mortimer 2004), displaced eastwards from Gondwana between 83 and 54 Ma by mid-ocean spreading of the Tasman Sea. Now largely submerged, Zealandia protrudes above sea level as the islands of New Zealand, the Chatham Islands and New Caledonia. Though formed from continental crust, these islands result from local tectonic activity but have geological histories extending back prior to 40 Ma, in marked contrast to the short-lived volcanic islands. Other islands are likely to have existed in the past, and terrestrial plant fossils dredged from a seamount west of the Three Kings Ridge suggest that a large island existed there between 38 and 21 Ma (Meffre *et al*. 2006).

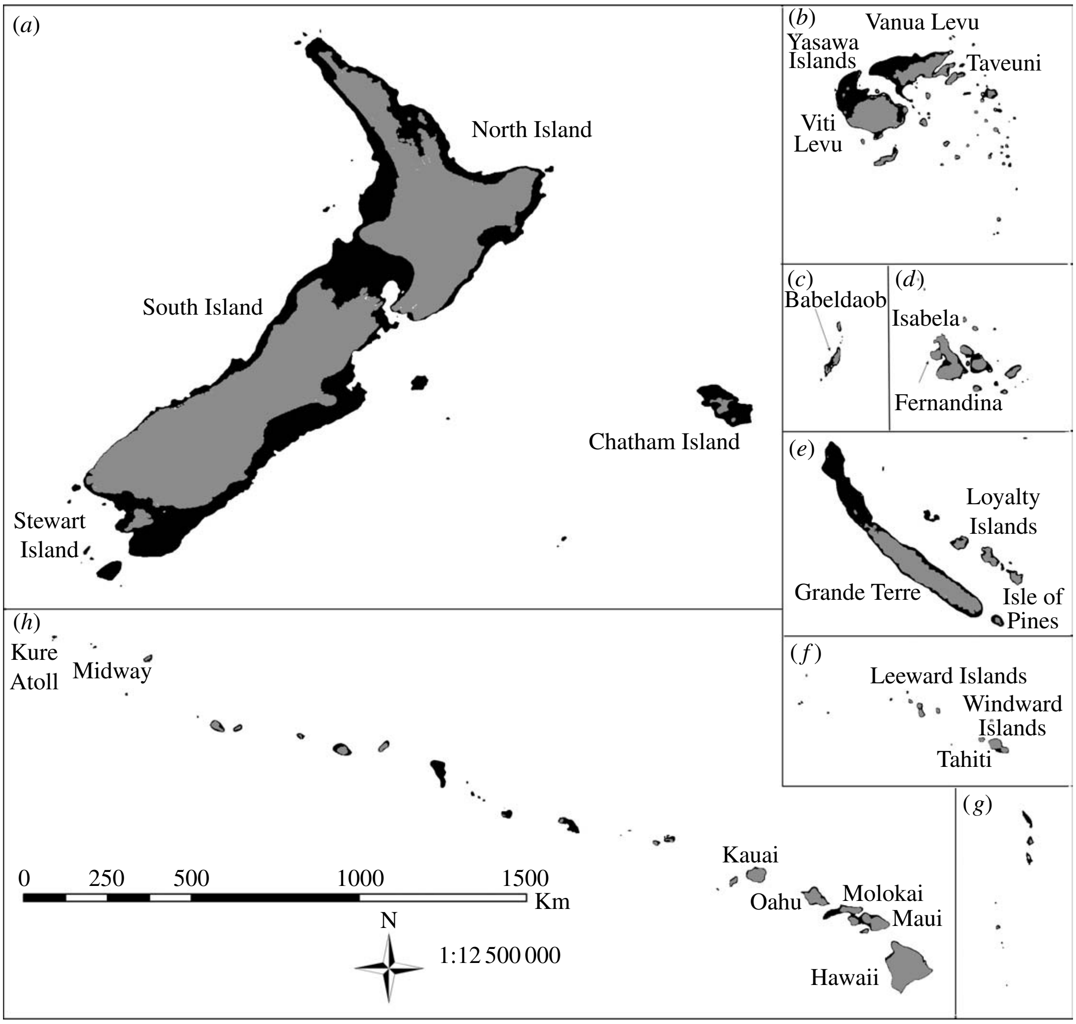

Figure 2. Maps of each selected archipelago exemplar (this volume) drawn to scale. Relative size (grey; present land) and probable area during the Pleistocene Last Glacial Maximum based on 120 m depth bathymetric contour (black). (*a*) New Zealand, (*b*) Fiji, (*c*) Palau, (*d*) Galápagos, (*e*) New Caledonia, (*f*) Society Islands, (*g*) Bonin Islands, (*h*) Hawaiian Islands, (*i*) Japan and (*j*) Solomon Islands.

(e) Formation of island arcs on the Pacific margins, mostly due to subduction of the Pacific Plate and its now extinct contemporaries (in particular the Kula and Izanagi plates). As the lithosphere descends into the mantle, the upper layers release water that reduces the melting point of the surrounding mantle. Melting generates magma that rises into the overlying lithosphere and may be emplaced either as intrusions or be extruded from the crust as volcanics. The formation of volcanic island arcs is most noticeable in the northern and western Pacific, creating the Aleutian, Japan–Kuril and Tonga–Kermadec arcs. In the western Aleutians, little or no convergence is occurring today so the margin is dominantly strike–slip, but in the central and eastern Aleutians convergence reached 100 mm yr^{-1} between 35 and 55 Ma, decreasing to 40–70 mm yr^{-1} since 15 Ma (Sdrolias & Müller 2006). The Japan–Kuril Arc has been a subduction zone since at least Cretaceous times with convergence ranging from 100 mm yr^{-1} in the south to 80 mm yr^{-1} in the north (Sdrloias & Müller 2006). The Tonga–Kermadec subduction zone was probably initiated *ca* 45 Ma and shows high convergence rates of 130–240 mm yr^{-1} in its northern half, declining to 20–60 mm yr^{-1} in its southern half (Sdrolias & Müller 2006).

Occasional large outpourings of basaltic lavas in submarine settings have also contributed to the plate geology (Kroenke 1996). The most notable is the Ontong Java Plateau, north of the Solomon Islands, which at 60 million km^3 forms the most voluminous large igneous province on Earth (Gladczenko *et al.* 1997). Such outpourings may represent the head of deep mantle plumes that initiate volcanic chains of islands. Although these large igneous provinces do not directly form islands, where they are carried to a plate boundary, they may indirectly lead to uplift and island formation on an adjoining plate.

In addition to the geological construction of islands, a number of other independent processes influence island formation—especially those associated with globally fluctuating sea levels during Quaternary time. During cool periods (the glaciations), when polar ice caps expanded, and ice sheets developed primarily on Northern Hemisphere continents, global sea level dropped to approximately 120 m below the present (Fairbanks 1989). This had the effect of linking islands

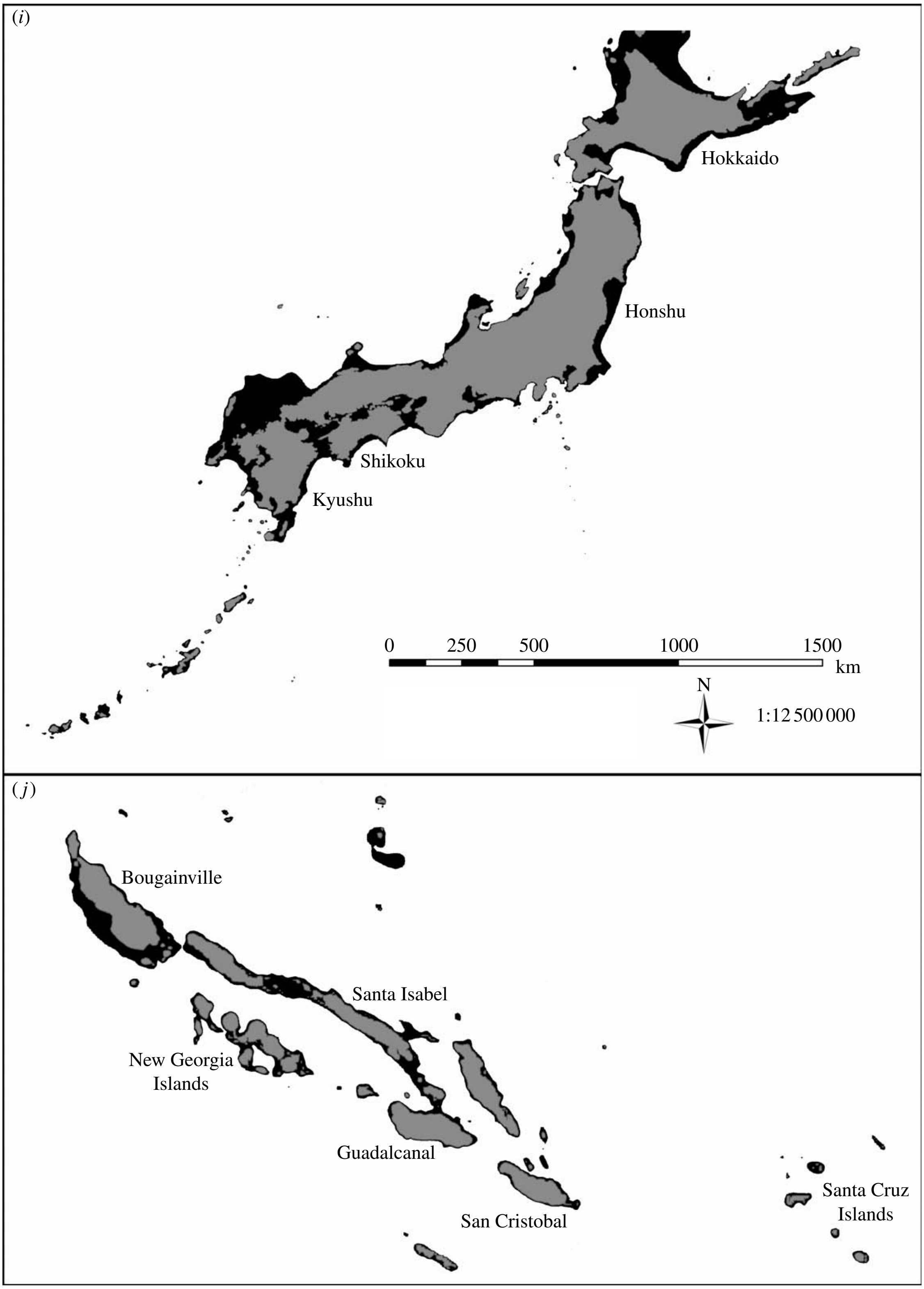

Figure 2. (*Continued.*)

that were surrounded by shallow seas (figure 2). For example, all the islands of the Chatham group would have formed a single large terrestrial island at least thrice larger than the present. During warm periods (the interglacials), when sea level was approximately similar in height to the present, the lower land of the 'glacial islands' became inundated by the sea (marine transgression) to isolate the higher altitude land as multiple, smaller islands. Given that glacial phases were more protracted (41–100 Ka) than interglacials (approx. 20 Ka), low glacial sea levels are likely to have significantly influenced island biodiversity.

During interglacials and in the presence of copious supplies of sediment, construction of land bridges (often in the form of sand bars or aeolian sand dunes) may have joined islands together. A classic example is the Aupori Peninsula in Northland, New Zealand, which formed in the last 0.5 Ma and provided a persistent terrestrial link between the former island at the northern tip of New Zealand to the land farther south. Volcanism can also create land bridges between islands, as has occurred on the Auckland isthmus (New Zealand), where basaltic lavas joined the former island of Northland to the North Island in Late Quaternary time. Tectonic uplift also has the potential to join, separate (e.g. around Scinde Island in the 1931 Napier earthquake, New Zealand) or enlarge existing islands, more particularly in the plate boundary regions of the Pacific. Tectonic subsidence can cause total submergence of islands and thus cause local biotic extinctions, e.g. the Louisville Seamounts.

4. THE PACIFIC LINEAR VOLCANIC CHAINS

A dominant feature on all bathymetric maps of the Pacific Ocean is the linear arrangement of volcanic edifices (volcanic islands and submarine seamounts) ascribed to either plumes derived from the Earth's core/mantle boundary region ('primary plumes'), or plumes derived from the upper/lower mantle transition zone, piercing their way upwards through the moving lithosphere, or to tensional fractures within the lithospheric plate allowing the release of magma from the asthenosphere/lithosphere boundary region. If the hot spot hypothesis is to be preferred, then one must also factor in that hot spots may move or be stationary over geological time (Gordon & Cape 1981; Tarduno *et al.* 2003, 2007; Parés & Moore 2005).

The Pacific plate is usually considered to host 14 hot spots, but Courtillot *et al.* (2003) argued only three are deep-source primary plumes and only two of the three show clear evidence of their beginnings in an oceanic plateau.[10] These are: the Hawaii—Emperor Seamount Chain (although this chain's origins are lost by subduction in the Aleutian Trench); Louisville; and Easter Island.

Of the other volcanic island chains and seamounts, another two, the Marquesas and Galápagos, have insufficient information yet available to confirm a deep mantle ('primary plume') origin.

A further eight show short tracks (less than 40 Ma) that cannot be traced to an ocean plateau. These are interpreted as hot spots related to plumes originating from the base of the upper mantle. These comprise: Foundation; Macdonald (in the Cook–Austral chain); Pitcairn–Gambier; Rarotonga; Rurutu or Raivavae; Arago; Society; and Samoa.

Recently, Wessel & Kroenke (2007) have drawn attention to the fact that all of these lineaments appear restricted to the Samoa–French Polynesia region of the Pacific, and suggest that they have resulted from a long-lasting tensile stress field[11] in this part of the Pacific plate caused by the subducting lithosphere ('slab pull') along the western Pacific margin.

Three other questionable Pacific hot spots, Kodiak–Bowie, Cobb and Caroline, are in such close proximity to subduction zones that their prior history has been lost. Older hot spot trails in the western Pacific probably date from 70 to 140 Ma, some of which may once have been islands that have since subsided.

The Society Islands, Galápagos Islands and Hawaiian Islands are discussed in detail later in this paper, hence here only the history of the other linear volcanic chains is briefly discussed. The Louisville Ridge is a 4300 km submarine chain of seamounts extending from the Louisville hot spot identified at 51°30′ S, 141°00′ W (Raymond *et al.* 2000) to the Tonga–Kermadec Trench, where it is being subducted to the west. The youngest volcanics are dated from a seamount at 50°26′ S, 139°09′ W as 1.11 Ma (Koppers *et al.* 2004). The chain comprises over 60 volcanoes spaced less than 100 km apart. Significant from the perspective of evolution on islands, there is evidence that 40 of these seamounts (all formed prior to 12 Ma) grew above sea level and were subsequently planed by marine erosion to form coral-free guyots.[12] During the Cenozoic, volcano building was remarkably constant at $3–4\times10^3\ km^3\ Myr^{-1}$, until 20 Ma when the hot spot activity declined sharply (Lonsdale 1988). Recent palaeomagnetic studies have cast doubt on this hot spot being responsible for the formation of the Ontong Java Plateau in the Early Cretaceous (Riisager *et al.* 2003).

The Easter hot spot lies close to Rapa Nui (Easter Island), an oceanic intraplate volcanic island located approximately 350 km east of the fast spreading EPR. Their close proximity may have initiated the large-scale rift propagation that created the Easter Island microplate *ca* 4.5 Ma (Hagen *et al.* 1990). There is a suggestion that this hot spot has generated two volcanic chains, one on each of the Pacific and Nazca plates. The principal volcanic chain (Easter Seamount Chain) extends from Pukao Seamount, through Easter Island and the Sala y Gomez and Nazca Ridges to the islands of San Felix and San Ambrosio, off the coast of Chile (Bonatti *et al.* 1977). A similar alignment to the west on the Pacific plate may be responsible for the Tuamotu Islands (Clouard & Bonneville 2001) and possibly the Mid-Pacific Mountains.

The Marquesas Archipelago is the northernmost linear volcanic chain in French Polynesia. It comprises eight main islands and a few islets and seamounts formed between 5.5 and 0.4 Ma (Legendre *et al.* 2006). Crough & Jarrard (1981) considered the Marquesas lineament to extend farther eastwards to the Line Islands, thus representing the track of a Marquesas hot spot over 40 Ma. Clouard & Bonneville (2001) went further and proposed that the hot spot originated at the time of the emplacement of the Hess Ridge and Shatsky Rise. The most recent interpretation of the chain suggests that it was formed by Pacific plate movement of $105\ mm\ yr^{-1}$ in a northeast direction over a shallow mantle plume (Legendre *et al.* 2006).

The Foundation Seamounts form a 1900 km long northeast–southwest-trending chain north of the Austral Islands. The Foundation hot spot near 38° S, 111° W (Mammerickx 1992; Devey *et al.* 1997) is close to the Pacific–Antarctic Ridge. Volcano ages increase progressively westwards to 21 Ma (O'Connor *et al.* 1998) and probably then follow the Ngatemato chain dated at 30 Ma (Clouard & Bonneville 2001). It appears some of the seamounts are currently so shallow that they may have once emerged above the sea surface. Each of these volcanic edifices probably represents *ca* 1 million years of construction (Maia *et al.* 2005).

The Macdonald hot spot is located in the southeastern Austral Islands where at 29° S, 140° W, the active Macdonald Seamount with a basal diameter of 45 km rises 3760 m above the surrounding sea floor. By 1987, it was just 39 m below sea level (Stoffers *et al.* 1989). The Macdonald hot spot trail can be traced back to Mangaia in the Cook Islands, which is dated at 19 Ma.

The Pitcairn–Gambier volcanic chain appears to have originated from hot spot volcanism on the Farallon plate at *ca* 25 Ma but the Pitcairn hot spot is now located on the Pacific plate, approximately 80 km east of Pitcairn Island. One of the submarine peaks, named Adams, reaches to within 20 m of sea level and is likely to have been exposed as an island during the Last Glacial Maximum. The hot spot appears to have been responsible for the islands extending from Pitcairn to the Gambier Islands in French Polynesia, to Mururoa, and then the Duke of Gloucester Islands in the Tuamotu Archipelago (Hekinian *et al.* 2003).

The Rarotonga hot spot is probably young, apparently having produced only Rarotonga Island (Clouard & Bonneville 2001), dated between 1.1 and 2.3 Ma (Thompson *et al.* 1998).

A volcanic chain extending from Raivavae Island (dated at 6.5 Ma) to Rurutu Island and nearby seamounts (dated at 12 Ma) in the Austral Islands has been recognized as the 'Raivavae hot spot track' (Bonneville *et al.* 2006) or Rurutu hot spot (Clouard & Bonneville 2001). A separate track nearby, extending from Arago Seamount (dated at 0.2 Ma) southeast of Rurutu Island to Atiu Island (dated at 8 Ma) has been recognized as the 'Arago hot spot track' (Bonneville *et al.* 2006).

The Samoa hot spot seems to have existed for at least 23 million years and was responsible for the five major islands of Samoa as well as a lineament of submarine seamounts and banks extending 1300 km west from Savai'i (Natland 1980; Clouard & Bonneville 2001; Hart *et al.* 2004). An active submarine volcano, Vailulu'u Seamount, approximately 30 km east of Ta'u, appears to mark the current location of the Samoa hot spot. Ta'u has a youthful uneroded shield morphology, while the islands to the west (Savai'i dated at up to 5.21 Ma, Koppers *et al.* 2006; Upolu dated at 3.2–1.4 Ma; and Tutuila dated at 1.53–1.0 Ma, Natland 2003) demonstrate various stages of advanced erosion (Hart *et al.* 2004). An isolated volcano (Machias Seamount), south of Upolu and adjacent to the Tonga Trench, has been dated at 0.97 Ma (Hawkins & Natland 1975). Samples dredged from the summit appear to have been deposited in a subaerial environment, suggesting that this structure has subsided approximately 700 m in less than 1 million years (Natland 2003).

5. EXEMPLARS OF ARCHIPELAGO GEOLOGY

This volume brings together papers on the biology of 10 archipelagos in the Pacific and we discuss the geology of these in further detail. These archipelagos comprise the full range of geophysical attributes and geological histories observed in the Pacific. The diversity in size, position, island number and mode of the formation of archipelagos across the Pacific Ocean is extraordinary. For instance, the largest, Japan and New Zealand, are each between six and four times bigger than all the other islands combined (table 1). Japan and New Zealand are also further from the Equator (more than 30°) than all others and with the Bonin Islands are the only non-tropical islands. Of all the island groups, Hawaii reaches the greatest altitude, with the twin peaks of Mauna Kea at 4205 m and Mauna Loa at 4168 m. This and their tropical position provide these islands with the greatest range of climatic zones and associated vegetational communities. On the other hand, the considerable latitudinal range in Japan and New Zealand and their complex physiographies also provide a wide range of climate zones and vegetational biomes. With the exception of Japan and New Zealand, freshwater lakes are few and very small on most of the islands; the exception is the large brackish water lake on Rennell Island (table 1). In Palau, some lakes in the karst landscapes close to the sea actually contain salt water.

With approximately 3000 islands, the Japanese archipelago outnumbers all the other exemplar archipelagos, with Fiji comprising the second largest number of islands (table 1). Japan has the longest coastline at approximately 30 000 km, followed by New Zealand at 18 000 km. Apart from these, it is the Solomon archipelago that has the next longest coastline (approx. 6000 km). Some archipelagos comprise islands spread over wide tracts of the Pacific Ocean, particularly where territorial claims are exercised over remote islands (such as Rotuma in Fiji, Kure Atoll in Hawaii and Minami-jima in Japan). Of these exemplars, the Society Islands at 5800 km from Australia are the most distant from a continental landmass (table 1).

(a) *Japan*

Japan has four main islands, from Hokkaido in the north, to Honshu, Shikoku and Kyushu in the south (sometimes called the 'Home Islands' see figure 2), and over 6000 smaller islands spread across the northwest Pacific. It is the largest of the Pacific archipelagos, extending 3500 km. The country is mountainous (73%) and dominated by two chains of andesitic stratovolcanoes, including Mt Fuji (Fujisan), the highest peak at 3776 m. Very little flat area exists and most hill slopes are cultivated for food production. Rivers are generally short with steep channel slopes until they deposit their sediment on the coastal alluvial plains. The climate ranges from cool temperate in the north to subtropical in the south, but it is generally rainy with high humidity. Japan's territorial waters extend over 3000 km^2 of ocean, with the remotest island, Minami Torishima, located 1848 km southeast of Tokyo.

The geological history of Japan has been dominated by subduction of the Pacific plate from the east beneath the Eurasian plate to the west. In the last 20 million years, the Philippine Sea (PS) Plate advanced northwards to form the plate boundary and began subducting northwestwards beneath southern Japan. Only within the last 15 million years has sea floor spreading

Table 1. Summary of key physiogeographic features of ten exemplar archipelagos in the Pacific Ocean.

	Japan[a]	Bonin	Palau	New Caledonia	Fiji	New Zealand	Society	Galápagos	Hawaii	Solomons[b]
number of islands	6773	27	300	36	332[c]	74[d]	14	19[e]	21[f]	~1000
area of land (km^2)	377 751	84	488	19 060	18 270	268 690	1685	7880	16 641	28 450
latitudinal range (degrees)	24–46N	26–28N	03–09N	18–23S	12–22S	34–47S[g]	15–19S	02N–02S	18–29N	05–11S
longitudinal range (degrees)	123–147E	142–143E	131–135E	162–172E	176E–178W	166E–176W	148–155W	89–92W	154–179W	154–162W
latitudinal range (km)	2500	222	667	556	1113	1447	445	445	1224	663
minimum distance from equator (km)	2670	2894	334	2004	1335	3780	1670	0	2004	552
maximum altitude (m)	3776	450	242	1628	1324	3754	2241	1707	4205	2715
total coastline (km)	29 470	281	1519	2254	1129	18 000	2525	1667	1693	6000
largest natural freshwater lake (km^2)	670	<1	5	4	2	606	<1	<1	<1	155[h]
mean annual temperature (°C)	10–20	23	27	23	25	10–16	21	24	24	27
nearest continent	Asia	Asia	Australia	Australia	Australia	Australia	Australia	South America	North America	Australia
distance to the nearest continent (km)	170[i]	1490	2500	1220	2630	1520	5800	920	4100	1488

[a] excluding Bonin Islands.
[b] including Bougainville.
[c] 106 are inhabited; plus 512 islets.
[d] excluding islets.
[e] plus 42 islets.
[f] plus 19 islets.
[g] excludes Kermadec and Subantarctic islands.
[h] brackish.
[i] to Kyushu.

in the Sea of Japan separated Japan from the Eurasian plate. Recent GPS measurements have suggested that the main islands of Japan are sited on two independent plates separated from the Eurasian plate. The first is the Amur (or Amurian) plate extending from Lake Baikal in the west to southern Honshu and the islands of Kyushu and Shikoku in the east. Farther to the east and north is the Okhotsk plate (formerly considered part of the North American plate), embracing northern Honshu, Hokkaido, the Sakhalin Peninsula and Kamchatka (Bird 2003). The Ryukyu Islands to the south of Kyushu are now regarded as being on a separate, narrow miniplate—the Okinawa plate.

The older sedimentary and metamorphic basement rocks of Japan lie to the west of southern Honshu. Since Mesozoic times, the Japanese arc has progressively increased in width by the addition of accretionary complexes[13] above the subduction zone. Concomitant granitic plutonism accompanied this process from Cretaceous to Miocene times. Currently, the country is subject to frequent earthquakes (approx. 1500 felt earthquakes each year) and tsunamis. Of the 200 volcanoes in Japan, 60 are active.

(b) *Izu-Bonin-Volcano Island*

The Izu-Bonin-Volcano Island arc (figure 2) is an archipelago extending for approximately 1000 km south of Tokyo Bay. The Bonin or Ogasawara-shoto Islands are approximately 800 km from the coast of Japan and cover 104 km^2 (table 1). Most of the Bonin Islands have steep shorelines, often with 50–100 m high sea cliffs, but some of the islands are fringed with coral reefs and a number of beaches. Only two of 27 islands are inhabited—Chichijima-rettō and Hahajima-rettō.

Between 50 and 40 Ma, much of a now extinct North New Guinea (NNG) plate was consumed by a subduction zone along its northern margin, as well as by the Eurasian plate to the northwest (Hall 1997). A final result of this process was the subduction of the NNG–Pacific ridge, which caused massive extension in the Izu-Bonin–Mariana forearc, and initiation of the subduction system to its east (Sdrolias & Müller 2006). The resultant volcanics that form the Bonin Islands of today are so distinctive geochemically; being enriched in magnesium, chromium and silicon, they are named boninites.

By 40 Ma, the Pacific plate was being subducted westwards beneath the Bonin Islands, and further Bonin Island volcanics were then formed along the northern boundary of the PS plate. By 30 Ma, Bonin Island volcanism ceased with rifting beginning between the Izu-Bonin arc and the Kyushu–Palau ridge to the west, leading to opening of the Parece Vela basin that continued until 15 Ma (Sdrolias *et al*. 2004).

(c) *Palau (Belau)*

The Palau archipelago comprises 12 inhabited islands and over 700 small islands and islets located 750 km east of The Philippines and north of the Equator (figure 2). They total 488 km^2 land area and stretch over 700 km of ocean. There is a great diversity of landforms and substrates for such a small area. The archipelago can be subdivided into four geological types. The volcanic islands Babeldaob, Ngerekebesang (also known as Arakabesan, Meiuns or Meyungs), Ngemelachel (or Malakal) and the western part of Oreor (or Koror) comprise most of the land area. Seven reefs and atoll islands are located north and northeast of Belilou (or Peleliu). The southwestern islands are a combination of low platform islands and atolls, while the 300 central and southern islands form steep coralline limestone rock islands. Substantial phosphate reserves were mined during the German and Japanese administrations.

Throughout the year the climate is hot, tropical and humid. Rainfall averages 3700 mm per year, ranging from 3150 to 4400 mm, with February to April the driest and the mean daily temperature is 27°C with a mean diurnal range of 7°C (Bureau of Agriculture 2003). Freshwater is limited, with only one sustained stream flow on the large volcanic island of Babeldaob. Freshwater lenses are also found beneath some of the atolls (www.sopac.org). Palau has a flora that is richer than any other area of Micronesia, with a large number of endemic species, the combined result of proximity to Asia and isolation (Canfield 1981).

The main island of Babeldaob (78% of the total land area) comprises in part old-weathered continental rocks that have been the source of some gold. Between 50 and 27 Ma, the Kyushu–Palau Ridge was a very active volcanic arc (Sdrolias *et al*. 2004) formed above the westward subducting Pacific plate. High-magnesium boninites and island arc tholeites were produced during Late Eocene time by rapid magma production. The Kyushu–Palau Ridge subsequently split off from the Bonin Islands *ca* 30 Ma, by sea-floor spreading of the Parece Vela basin. In Palau, the volcanics are dated between 37.7 and 20.1 Ma (Kobayashi 2000). They are in turn overlain by Mid-Oligocene limestone oozes, indicative of subsequent subsidence. Rotation of the PS plate gradually led to a convergent plate boundary with the Caroline plate to the southeast, with subsequent uplift of the islands raising Miocene limestones to 220 m asl, forming a limestone cap (Canfield 1981; Kobayashi 2000) with surrounding raised coral platforms and reefs. A 4000-year-old limestone on Oreor has been subsequently uplifted 2 m, i.e. current average rate of uplift is 0.5 mm yr^{-1} (Kobayashi 2000).

(d) *The Solomon archipelago*

This archipelago forms a double chain of islands aligned northwest–southeast across 1300 km of the western Pacific Ocean, comprising the nation of the Solomon Islands and the adjoining islands of Bougainville and Buka in Papua New Guinea. There are seven major Solomon Islands comprising one alignment to the south of Vella Lavella, New Georgia, Guadalcanal and Makira (also known as San Cristobal) and a second alignment to the north of Choiseul, Santa Isabel and Malaita. The remote island of Rennell located approximately 200 km south of Guadalcanal is a raised coral limestone with a large brackish water lake (Te Nggano), reputedly the largest freshwater lake in the 'insular' Pacific. About a further 340 smaller islands are populated. The islands are bounded to the northeast and south by deep oceanic trenches, while to the north is the voluminous submarine Ontong Java Plateau. Most of the inland terrain of the Solomon archipelago is covered with dense tropical rainforest. The highest points are Mt Makarakomburu on Guadalcanal at 2447 m in the Solomon Islands and Mt Balbi at 2715 m on Bougainville Island. The annual temperature is fairly consistent between 28 and 30°C; the mean annual rainfall at the capital Honiara is 2800 mm.

The geological history of the Solomon arc is complex having formed at the southwestern boundary of the Pacific plate in contact with the Australian plate, except at the western end where the Woodlark and Solomon Sea plates are subducted to the north. In Cretaceous times, the Pacific plate was subducted southwards in this region (Schellart *et al.* 2006). From 55 to 40 Ma, the Australian plate was also being subducted northwards (Kroenke 1996). Approximately 25–20 Ma, the Ontong Java Plateau on the Pacific plate first made contact with the arc and since 4 Ma has forcefully collided, leading to obduction (Petterson *et al.* 1999).

The geology of the Solomons is a collage of geological units that have been thrust together during this complex tectonic history. The north coast of Santa Isabel, and all of Malaita and Ulawa, represents obducted basalt flows and sedimentary cover, geochemically identical to the Ontong Java Plateau (Petterson *et al.* 1999). A second geological unit comprising basaltic flows, ultrabasic intrusions and overlying sediments, typical of a mid-ocean spreading centre origin, forms the islands of Choiseul and Guadalcanal (Petterson *et al.* 1999). A third unit showing hybrid characteristics of the above two occurs on San Cristobal. Subsequent arc volcanics were emplaced in two discrete phases. The first phase lasted from 62 to 24 Ma, and the second is responsible for volcanic activity from 7 Ma to the present day (Petterson *et al.* 1999). Currently, there are five active volcanoes in the Solomon arc: the shallow submarine volcano Kavachi, the stratovolcano island of Savo, Tinakula in the Santa Cruz Islands, the fumarolic volcanic island of Simbo and Bagana volcano on Bougainville.

Construction of the islands of Bougainville and its neighbour Buka began as a submarine volcanic pile *ca* 45 Ma (Davies 2005). It is not clear when the islands became emergent, but a second construction phase began *ca* 10 Ma when subduction commenced from the south. Within the older volcanic rocks are coarse-grained intrusives of granodiorite and diorite, which contain low-grade copper, gold and silver mineralization (Davies 2005). The Panguna copper mine, with proven ore reserves in excess of 944 million tonnes, was worked between 1972 and 1988.

(e) *Fiji*

The Fiji group of islands has a total land area of 18 270 km^2 (table 1). Of the 332 islands in the group, approximately 110 are inhabited, with an additional 522 uninhabited islets (figure 2). The two largest islands are Viti Levu (10 642 km^2 and 57% of the nation's land area), where the capital Suva with a population of 200 000 is located, and Vanua Levu (5807 km^2). Both islands are mountainous with peaks rising to 1300 m. This results in a wet windward southeastern side that was originally covered in dense tropical rainforest, and a dry-season leeward western side, highly favourable for crops such as sugarcane. The third largest island is Taveuni, renowned for its rich volcanic soils. Approximately 1000 coral reefs surround the islands.

The geological history of Fiji is complex owing to its proximity to the Australian–Pacific plate boundary. The oldest rocks in Fiji are island-arc volcanics of Late Eocene age formed by westwards subduction of the Pacific plate beneath the Australian plate along the Vitiaz arc. These strata are now found in western Viti Levu. Their subsequent uplift allowed deposition of shallow-water limestones around them. After 28 Ma relative movement between the plates changed to a more oblique convergence and led to another major suite of volcanic islands and sedimentary basins. These strata are now exposed in southern Viti Levu, and in the Yasawa and Mamanuca Islands. To the north, coral reef limestones accumulated. During the Middle to Late Miocene, plutonic intrusions were emplaced and uplift produced the first significant land mass, mostly Viti Levu, at this time. Approximately 12 Ma, the configuration of the plate boundary changed with break-up of the Vitiaz Trench, and establishment of a major transform boundary to the north, the Fiji Fracture Zone. Approximately 10 Ma spreading was initiated in the North Fiji Basin, and Viti Levu became attached to the Pacific plate. Just before 7 Ma, Vanua Levu was formed and rotated clockwise as the Central North Fiji Basin triple junction developed (Kroenke 1996). Since 7 Ma Fiji has rotated anticlockwise with the opening up of the North Fiji Basin, and strike–slip faulting takes up the relative movement to the north of Vanua Levu. Taveuni (437 km^2) is a basaltic volcanic island with a northeast–southwest lineament of over 150 Quaternary vents along its 40 km length. At least 2.7 km^2 of magma has been erupted from over 100 events on Taveuni during the Holocene (Cronin & Neall 2001).

(f) *New Caledonia*

New Caledonia comprises the main island (Grande Terre), the Isle of Pines to the south, the Loyalty Islands to the east (Maré, Lifou, Tiga and Ouvéa) and the Belep Archipelago to the northwest (figure 2). Grand Terre is the third largest island in the southwest Pacific, after Papua New Guinea and New Zealand. All the islands comprise a total land area of 18 575 km^2 (table 1). Grande Terre is surrounded by a 1600 km long coral reef, forming the largest lagoon in the world (16 000 km^2). Proximity of the reef to the coast varies from a few kms to 65 km, the lagoon having an average depth of 40 m. A chain of mountains (Chaine Centrale) forms a backbone to Grand Terre, dividing the island into a humid east coast and a drier west coast. The highest peak is Mt Panié at 1629 m.

Grand Terre is a fragment of continental crust (Zealandia) that rifted northeastwards away from Gondwana in Late Cretaceous times, *ca* 65 Ma. What is now the current island reached its present position *ca* 50 Ma. It then attained its current geological configuration *ca* 45 Ma when the Loyalty Arc collided with what is now New Caledonia between 38 and 33 Ma (Schellart *et al.* 2006). This convergence culminated in obduction from the northeast, emplacing a 2 km thick sequence of ophiolites; sections of previous basaltic ocean floor and entrained sediments (Aitchison *et al.* 1995). Subsequent uplift and erosion has reduced this to approximately one-third (5500 km^2) of the total original land area. These rocks are high in iron and magnesium (ultramafics) and give rise to approximately 25 per cent of the world's known nickel resources and unusual red soils enriched in nickel, chromium, cobalt and manganese. The soils have had a profound influence on the flora that has adapted to these low nitrogen, phosphorus, potassium and calcium conditions (Jaffré *et al.* 1987). Recent provenance studies of the Goa N'Doro Formation have suggested that fluvial aggradation on an emergent landmass (New Caledonia) had begun by Oligocene time (Chardon & Chevillotte 2006). Subsequent Quaternary uplift has resulted in formation of the emergent atolls of the Loyalty Islands (Dubois *et al.* 1974, Paris 1981).

(g) *New Zealand*

New Zealand consists of two large islands, North Island (113 729 km^2) and South Island (150 437 km^2), and two smaller islands, Stewart Island (1680 km^2) and the Chatham Islands (970 km^2) (figure 2). There are approximately 700 other small islands and islets, mostly within 50 km of the main islands. The islands extend across 1500 km of ocean from 34° to 47° south (table 1). The highest point is Mt Cook (Aoraki) in the South Island at 3754 m. These islands developed on part of an extensive submarine plateau composed of continental crust (collectively named Zealandia) totalling 1.7 million km^2, an extension of the islands stretching from the Campbell Plateau to the southeast, and the Chatham Rise in the east, to the Challenger Plateau to the west and New Caledonia in the north.

Zealandia began to separate from Gondwana *ca* 90 Ma with the southernmost portions rotating anticlockwise about a pole of rotation to the north of the country (Mortimer 2004). Sea-floor spreading began *ca* 83 Ma and propagated northwards into the Coral Sea by 62 Ma (Gaina *et al.* 1998). Eventually, the northernmost parts separated to create the Tasman Sea, until 52 Ma when sea-floor spreading ceased (Schellart *et al.* 2006).

By this time, Zealandia was gradually subsiding, as a marine transgression covered much of the region that would later become New Zealand, depositing widespread limestone, calcareous sandstone and siltstone and greensand. Whether this submergence was complete is not known though it has long been argued that some islands persisted on which the Zealandian elements of New Zealand's flora and fauna survived (Fleming 1962). However, the direct geological evidence for persistent land is equivocal (Campbell & Hutching 2007; Landis *et al.* 2008)

Australia then began to separate from Antarctica by new sea floor spreading along the Indian–Antarctic Ridge. This eventually subjected the New Zealand region to major lateral and oblique compressive movements, beginning *ca* 30 Ma, that have sliced through the South Island to form the Australian–Pacific plate boundary (Kamp 1986; King 2000). This marks the beginning of the formation of New Zealand we know today. Approximately 470 km of dextral horizontal movement along the Alpine Fault is recorded over this time period. Accelerated oblique convergence between the Australian and Pacific plates then led to the subsequent deformation of the New Zealand region and the uplift of the Southern Alps since the onset of the Pliocene (5.3 Ma) (Kamp *et al.* 1989; Whitehouse & Pearce 1992). At the northern end of the South Island, numerous faults splay through Marlborough to link the plate boundary to the Hikurangi Trough, east of the North Island. Rotation of a miniplate forming the eastern North Island has probably been responsible for the active continental rifting of the Taupo Volcanic Zone (TVZ) and its attendant rhyolitic and andesitic volcanism in the last 2 million years (Wallace *et al.* 2004). Over 10 000 km^3 of dominantly rhyolitic magma has been erupted from the TVZ in the last 1 million years (Wilson *et al.* 1984).

Intraplate volcanism formed the Sub-Antarctic Islands to the south of mainland New Zealand (Quilty 2007), in the order: the Auckland Islands (25–11 Ma), followed by Campbell Island (11–6 Ma) and lastly the Antipodes (5 to less than 1 Ma) (Gamble & Morris 1989). The Chatham Islands have probably emerged in the last 4 million years by as yet unexplained tectonic uplift on the far eastern margin of Zealandia (Campbell *et al.* 1993).

(h) *Society Islands*

The Society Islands are the best known of the five island groups comprising French Polynesia. The islands are spread across 720 km of the Pacific Ocean and comprise two groups—the Windward and Leeward Islands. In total, the land area is 1680 km^2 with the main island of Tahiti (1042 km^2) being the largest (figure 2).

The Windward Islands in the east comprise Tahiti on which the main population centre of Papeete is located, Moorea (132 km^2), steep-sided Mehetia (all volcanic) and the atoll Tetiaroa. The Leeward Islands in the west comprise the mountainous volcanic islands of Ra'iatea (238 km^2), Taha'a (91 km^2), Huahine (74 km^2) and Bora Bora (22 km^2) and the atolls of Maupiti, Maiao, Maupihaa, Tupai, Manuae and Motu One. The Society Islands are thought to have originated from a hot spot located near the island of Mehetia, 110 km east of Tahiti (Devey *et al.* 1990). The chain exhibits an age progression to the west from Mehetia (less than 1 Ma) to Tahiti (0.6–1.2 Ma), Moorea (1.5–2.0 Ma) Bora Bora (3.1–3.5 Ma) and Maupiti (3.9–4.5 Ma) (Blais *et al.* 2002; Yamamoto *et al.* 2002; Uto *et al.* 2007), consistent with a plate motion of 110 mm yr^{-1}. The climate is tropical with a mean annual temperature of 26°C and mean annual rainfall from 1700 mm near sea level to 8000 mm around the high mountain peaks (www.oceandots.com).

(i) *Galápagos Islands*

The Galápagos Islands (or Archipelago of Colon) lie on the Equator, 1000 km west of South America. They comprise 13 main islands, and 48 smaller islands and islets forming a total land area of 7880 km^2 within a geographical area of 45 000 km^2 (figure 2). The largest island is Isabela, one of the two youngest, making up more than half the total land area at 4855 km^2. The summit of Volcan Wolf forms the highest point at 1701 m. With the exception of Isabela, all the other islands represent single shield volcanoes in various stages of marine erosion. Isabela comprises six shield volcanoes, Sierra Negra erupting in October 2005 (Geist *et al*. 2006).

The westernmost islands, Fernandina and Isabela, are regarded as representing the current location of the Galápagos hot spot. The oldest exposed rocks date from only 3 Ma, but seamounts on the Cocos Ridge on the Cocos plate, and the Carnegie Ridge on the Nazca plate, extend back to 14 Ma and may have been subaerial islands (Christie *et al*. 1992; Werner *et al*. 1999). Palaeomagnetic and geochemical data demonstrate a complex interaction over geological time between the hot spot and the nearby Cocos–Nazca spreading centre, with the hot spot varying from being to the south (19.5–14.5 Ma) to the north (14.5–12 Ma), to being coincident (11–12 Ma) (Werner *et al*. 2003). Furthermore, Hoernle *et al*. (2002) suggested igneous complexes along the Pacific margin of Costa Rica and Panama represent volcanics emplaced by the Galápagos hot spot between 20 and 71 Ma.

(j) *Hawaiian Islands*

The Hawaiian Islands are wholly volcanic, principally shield volcanoes formed over the Hawaiian hot spot as the Pacific plate moved progressively towards the WNW (see the electronic supplementary material and figure 2). The oldest of the Hawaiian Islands is at the western end of the chain, Kure Atoll formed *ca* 25 Ma. The islands become progressively larger and younger towards the ESE, where the youngest and still volcanically active island of Hawaii (10 458 km^2) exceeds the combined land area of all the other islands in the State (total for the State is 16 705 km^2). The island also hosts the two highest peaks in the archipelago—Mauna Kea (4205 m) and Mauna Loa (4169 m), with 68 per cent of the island above 610 m. The second largest and youngest is Maui (1888 km^2) dated between 1.32 and 0.75 Ma (McDougall 1964; Naughton *et al*. 1980). This is followed in age progression by Molokai (676 km^2) dated between 1.75 and 1.9 Ma (Naughton *et al*. 1980), Oahu (1574 km^2) dated between 2.6 and 3.7 Ma (Doell & Dalrymple 1973) and Kauai (1433 km^2) dated at 5.1 Ma (McDougall 1979). The coastline of all the islands sums to a length of 1207 km (University of Hawaii 1973).

The big island of Hawaii is the nearest existing island to the Hawaiian hot spot, considered to be in the vicinity of Loihi Seamount at 19° N, 155° W. The chain then extends northwestwards through the Hawaiian Islands to Midway, and at 32° N 173° E there is a pronounced change in direction to the north forming the Emperor Seamount Chain. With increasing distance from source, the 129 volcanoes of the chain gradually subside with their accompanying oceanic plate to depths of 2 km (electronic supplementary material). At the northern terminus of the Chain is Meiji, once comparable with Hawaii that is next to be subducted into the western Aleutian Trench, a total distance of 6000 km from Hawaii (Clague 1996). The age of the bend in the alignment of the volcanic chain is now well dated at 47 Ma (Sharp & Clague 2002; Tarduno 2007). Initially, this was interpreted as a change in the movement of the Pacific plate over the Hawaiian hot spot, but more recent palaeomagnetic information in particular has suggested that between 76 and 47 Ma, when the Emperor Seamount Chain was emplaced, the Hawaii hot spot moved southwards at more than 30 mm yr^{-1} across a WNW moving Pacific plate (Tarduno & Cottrell 1997; Tarduno *et al*. 2003). Since 47 Ma the hot spot appears to have remained relatively fixed with respect to the spin axis of the Earth (Tarduno *et al*. 2003; Sager *et al*. 2005), although contrary evidence suggests it moved at 32 mm yr^{-1} to the southeast during the past 43 million years (Pares & Moore 2005). However, many workers have pointed out that a number of key criteria for a hot spot are missing in Hawaii. These include no initial 'plume head' to the chain represented by flood basalts or oceanic plateaus (Campbell & Griffiths 1990, 1993); highly variable eruption rates along the chain (Robinson & Eakins 2006); no significant heat flow anomaly (von Herzen *et al*. 1989; Stein & Stein 1992, 1993; McNutt 2000); the melt originates from the shallow asthenosphere (Sisson in Foulger & Anderson 2006); and seismology has not detected a plume (Wolfe *et al*. 2002). Notwithstanding this evidence, the ages of the islands show a very regular time-progressive track. Alternatively, the chain may have originated by a propagating fracture controlled by the direction of regional stress, the fabric of the seafloor or stresses caused by previously erupted volcanoes (Hieronymus & Bercovici 1999).

6. ENVOI

The above overview demonstrates the dynamic and complex history involved in the formation of the Pacific islands. This complexity is still being unravelled from new marine exploration of the seabed. The discovery in the past 50 years of a wide array of subsided seamounts throughout the Pacific that once formed above sea level provides insight into the potential origins and age of island floras and faunas. The array of geological processes and resultant geochemistry and physiography is implicated in the development of complex floral and faunal histories on each island. Evidence from molecular techniques is beginning to illuminate the details of this complexity, and in some cases reveals aspects of island geology not preserved in the stratigraphic record, providing new lines of evidence for island geological histories.

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ENDNOTES

[1]Plate nomenclature follows Bird (2003).

[2]Subduction: the process by which one tectonic plate moves beneath another downwards into the mantle.

[3]Lithosphere: the outer rigid layer of the Earth comprising crust and upper mantle, which averages 100 km thick, but varies with age between 2 and 300 km thickness.
[4]Asthenosphere: the weak ductile layer in the upper mantle, upon which lithospheric plates move.
[5]Obduction: overthrusting of one lithospheric plate onto an adjoining plate at a convergent plate boundary. The obducted strata may be continental or oceanic in composition. In the latter case, sections of obducted ocean floor are called 'ophiolites'.
[6]Island arc: a (usually) curved chain of volcanic islands, usually positioned above a subduction zone.
[7]Strike–slip: a type of fault in which opposing movement of rock on either side is dominantly horizontal.
[8]Plume: upwelling of hot magma from the mantle.
[9]En echelon: short parallel or subparallel, closely-spaced, overlapping structural features in rock. Collectively, they may form a linear zone.
[10]Oceanic plateau: a large, undersea igneous province, the marine equivalent of continental flood basalts.
[11]Stress field: the distribution of stress (compression or tension) through a body of rock.
[12]Guyot: a seamount, the top of which has been planed by erosion.
[13]Accretionary complex: The accumulated strata at the boundary of an overriding plate above a subduction zone. The strata accumulate on the inner wall of an ocean trench as material on the upper part of the descending oceanic plate is scraped off or underthrust to accrete on or beneath the leading edge of the overriding plate.

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